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4

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Committed warming inferred from observations

Thorsten Mauritsen[1]* and Robert Pincus[2,3]

Due to the lifetime of CO_2, the thermal inertia of the oceans[1,2], and the temporary impacts of short-lived aerosols[3-5] and reactive greenhouse gases[6], the Earth's climate is not equilibrated with anthropogenic forcing. As a result, even if fossil-fuel emissions were to suddenly cease, some level of committed warming is expected due to past emissions as studied previously using climate models[6-11]. Here, we provide an observational-based quantification of this committed warming using the instrument record of global-mean warming[12], recently improved estimates of Earth's energy imbalance[13], and estimates of radiative forcing from the Fifth Assessment Report of the Intergovernmental Panel on Climate Change[14]. Compared with pre-industrial levels, we find a committed warming of 1.5 K (0.9–3.6, 5th–95th percentile) at equilibrium, and of 1.3 K (0.9–2.3) within this century. However, when assuming that ocean carbon uptake cancels remnant greenhouse gas-induced warming on centennial timescales, committed warming is reduced to 1.1 K (0.7–1.8). In the latter case there is a 13% risk that committed warming already exceeds the 1.5 K target set in Paris[15]. Regular updates of these observationally constrained committed warming estimates, although simplistic, can provide transparent guidance as uncertainty regarding transient climate sensitivity inevitably narrows[16] and the understanding of the limitations of the framework[11,17-21] is advanced.

Burning of fossil fuels elevates atmospheric concentrations of carbon dioxide (CO_2), alters atmospheric chemistry, and produces aerosol particles. Over the past century, warming by CO_2 and other greenhouse gases has exceeded cooling by aerosols and Earth's surface temperature has gradually increased. If fossil-fuel emissions were to cease instantaneously, anthropogenic aerosols would be washed out of the atmosphere in a matter of weeks but anthropogenic CO_2 would persist, equilibrating only across centuries to millennia. The long life of CO_2 and the large thermal inertia of the oceans imply that some amount of future warming is inevitable even in the unreasonably optimistic scenario of an abrupt halt to fossil-fuel emissions. Here we apply observational constraints within a simple linear energetic framework to estimate the magnitude of this committed warming due to past emissions.

We first estimate Earth's equilibrium climate sensitivity (ECS) and transient climate response (TCR) (Fig. 1 and Methods) from estimates of effective radiative forcing and observations of present-day energy imbalance and global-mean surface temperature, using an energy-balance model and treating uncertainty in each term using probability distributions[22]. Both sensitivities are defined according to the forcing by atmospheric CO_2 concentrations doubled from pre-industrial values; ECS is the warming that occurs when the deep oceans have equilibrated while TCR is the warming at the time of doubling, assuming forcing increasing linearly at a rate consistent with a 1% per year increase in CO_2 concentration. Our estimates of TCR are commensurate with previous estimates[23,24] and, despite reductions in the uncertainty of the Earth's energy imbalance[13], the uncertainty in inferred values of ECS remains large. Best estimates of TCR and ECS inferred from historical observations and an energy-balance framework are at the lower end of the assessed ranges (1.0–2.5 and 1.5–4.5 K, respectively) that consider a wider variety of evidence[25]. We return to this idea and its implications below.

These estimates of climate sensitivities can be used to infer the committed future warming on centennial (transient) and at multi-millennial (equilibrium) timescales by scaling the radiation imbalance and changes in forcing by the relevant climate sensitivity. The simplest estimate of equilibrium committed warming T_a is obtained by holding effective radiative forcing at present-day values and assuming that the warming balancing present-day energy imbalance ($\overline{Q}$) will be consistent with joint distributions of past response to forcing. The resulting increment is added linearly to the mean warming ($\overline{T}$) in the present-day (here 2005–2015) relative to a pre-industrial period (1850–1899):

$$T_a \approx \overline{T} + [\overline{Q} + \delta F] \frac{\mathrm{ECS}}{F_{2\times}} \qquad (1)$$

where $F_{2\times}$ is the radiative forcing from a doubling of CO_2. Here $\delta F \approx 0.2\ \mathrm{W\,m^{-2}}$ accounts for forcing by emissions from 2010 (the centre of the present-day period) to 2016, estimated from ref. 14 for the period 2000–2011 as about $0.033\ \mathrm{W\,m^{-2}\,yr^{-1}}$ to yield an up-to-date estimate of commitment. The result is case a in Fig. 2.

In the absence of fossil-fuel emissions, however, present-day aerosol forcing ($F_{\mathrm{aero}} \approx -0.9\ \mathrm{W\,m^{-2}}$ (-1.9 to -0.1)) will quickly vanish as anthropogenic aerosols are removed by precipitation and other processes (case b):

$$T_b \approx \overline{T} + [\overline{Q} + \delta F - F_{\mathrm{aero}}] \frac{\mathrm{ECS}}{F_{2\times}} \qquad (2)$$

where it is assumed that non-fossil-fuel anthropogenic aerosols, such as biomass burning, yield near-zero forcing[26]. Values of committed warming exceeding 2 K are far more likely under this assumption even though aerosol forcing does not dominate total forcing (Supplementary Table 1). The high values of equilibrium committed warming arise because, in the energy-balance model, the strict constraint on observed warming means that strong aerosol cooling is possible only if climate sensitivity is high. These estimates are nonetheless smaller than earlier estimates[5] of 2.4 K (1.4–4.3) applying equivalent assumptions but basing their ECS on climate models.

Although carbon dioxide is long-lived, other chemical species emitted by fossil-fuel burning, including methane (CH_4), nitrogen oxides (NO_x) and carbon monoxide (CO), impact the Earth's radiation balance[6] both directly and through chemical reactions affecting the concentrations of ozone, stratospheric water vapour, and each other. We estimate that short-lived climate forcers (SLCFs) taken together introduce a forcing of $0.36\ \mathrm{W\,m^{-2}}$ (0.17–0.56) (see

[1]Max Planck Institute for Meteorology, Bundesstrasse 53, 20146 Hamburg, Germany. [2]University of Colorado, Boulder, Colorado 80309, USA. [3]NOAA Earth System Research Lab, Physical Sciences Division, Boulder, Colorado 80305, USA. *e-mail: thorsten.mauritsen@mpimet.mpg.de

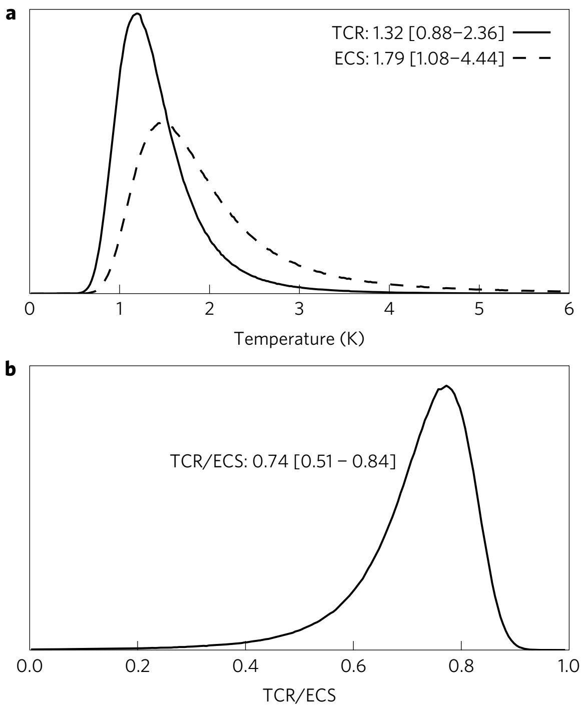

Figure 1 | Probabilities of transient climate response (TCR) and equilibrium climate sensitivity (ECS). a, Probabilities of TCR (solid) and ECS (dashed) inferred on the basis of observed warming, estimates of historical radiative forcing and observations of present-day energy imbalance. **b**, The ratio of the quantities in **a**, which is roughly equivalent to the proportion of long-term warming realized on centennial timescales. Displayed numbers are the median and 5th–95th percentiles of each distribution.

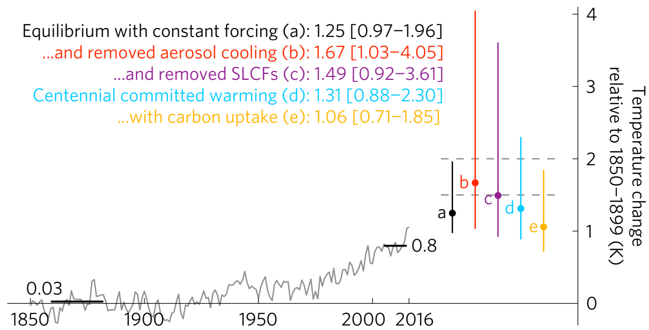

Figure 2 | Estimates of committed warming under five different sets of assumptions. Cases a (black) and b (red) are the equilibria with and without aerosol cooling, whereas case c (purple) includes the effect of removing short-lived climate forcers. Cases d (blue) and e (orange) are scaled with the transient climate response representative of warming within this century. Case d is otherwise equivalent to case c. In case e, it is assumed that carbon uptake on the centennial timescale cancels the remnant warming due to imbalance with past forcing. Displayed numbers are the median and 5th–95th percentiles of the respective distribution. Also shown in grey is the instrumental temperature record[12] for 1850–2016, and black horizontal lines indicate the reference periods used to estimate TCR and ECS (Fig. 1). The dashed horizontal lines indicate 1.5 and 2.0 K. All temperatures are relative to the 1850–1899 mean, which is here taken to be the pre-industrial reference temperature (Methods).

Methods). The loss of this net warming effect parallels the impact of reduced aerosol cooling:

$$T_c \approx T + [Q + \delta F - F_{aero} - F_{SLCF}]\frac{ECS}{F_{2\times}} \qquad (3)$$

The result is a slight reduction in equilibrium committed warming (Fig. 2 case c).

Estimating the amount of warming to be realized in the current century requires accounting for the multiple timescales of equilibration in the climate system. These timescales—in an idealized view, a yearly to decadal timescale associated with equilibration of the atmosphere, upper soil and ocean mixed layer, and a centennial to millennial timescale associated with the overturning of the deep oceans—imply that the temporal response of surface temperature is sensitive to the history of the applied forcing. An abruptly applied positive forcing, such as that arising from the cessation of anthropogenic aerosol emissions ($-F_{aero}$), is associated primarily with a fast warming near the surface, followed by slow warming, while equilibration with remnant planetary energy imbalance due to past forcing (Q) involves mainly a slow warming of the deep oceans. The fraction of equilibrium warming on centennial timescales may be estimated using ocean models of varying complexity[27] but these are poorly constrained by observations. Instead, we assume that the centennial response to present-day forcing will be consistent with the response to historical forcing, and so approximate centennial commitment using the observationally determined TCR

$$T_d \approx T + [Q + \delta F - F_{aero} - F_{SLCF}]\frac{TCR}{F_{2\times}} \qquad (4)$$

resulting in lower commitments than at equilibrium (Fig. 2). Large values of centennial-scale committed warming are unlikely because the high ECS values leading to large T_c are also associated with smaller ratios of TCR/ECS (Fig. 1b), such that the equilibrium commitment takes more time to be realized if sensitivity is high[28]. The approximate scaling of centennial warming by TCR is consistent with an elaborated energy-balance model with a two-layer ocean (see Methods). When tuned to be consistent with inferred ECS, TCR and Q, and driven by the time history of forcing until 2010, followed by forcing evolution consistent with case d until 2100 (Supplementary Fig. 1), this model produces committed centennial warming in agreement with the estimates using equation (4).

To this point we have assumed that atmospheric CO_2 concentrations are constant after fossil-fuel emissions cease. This is unlikely to hold because the oceans absorb carbon as well as heat from the atmosphere. Although the magnitude of ocean carbon uptake on centennial timescales is uncertain it must act to lower committed warming relative to fixed CO_2 concentrations (cases a–d). An estimate of this effect is obtained by idealizing the finding that in Earth system models temperatures stay approximately constant for one or more centuries after carbon emissions cease[7–11], suggesting that remnant warming is approximately cancelled by declining forcing due to oceanic carbon uptake. The approximation is consistent with the mixing process timescale for carbon into the deep ocean being the same as for heat and carbon uptake being approximately linear in CO_2 concentration. A crude representation of this cancellation is to neglect further warming from energy imbalance Q:

$$T_e \approx T + [\delta F - F_{aero} - F_{SLCF}]\frac{TCR}{F_{2\times}} \qquad (5)$$

which yields a reduction in estimated median committed warming of 0.2–0.3 degrees (Fig. 2, case e). This scenario was examined in an Earth system model by the authors of ref. 6 who found a fast decadal warming to +0.4 K followed by a slow decline in temperature to around +0.2 K over that at the time when emissions were stopped, in close agreement with the median estimate of +0.26 K found here.

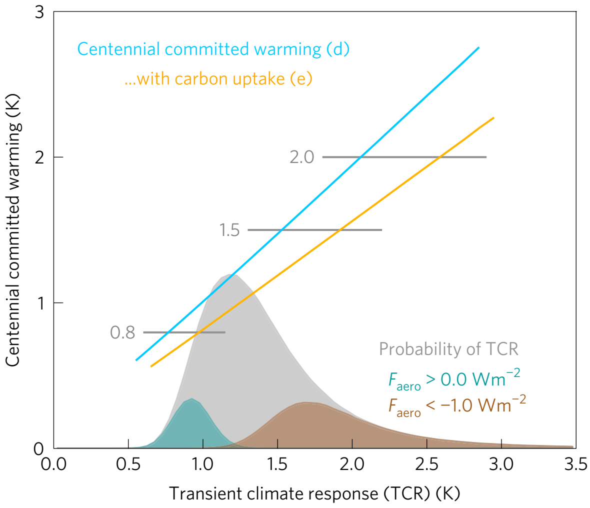

Figure 3 | Commitment as a function of transient climate response. Blue and orange lines show binned median commitment for cases d and e, respectively, as defined in Fig. 2. The shaded area shows the probability of TCR, repeated from Fig. 1, with overlaid green shading for cases with positive aerosol forcing and red shading for cases with aerosol forcing below $-1.0\ \mathrm{W\,m^{-2}}$. The warming in 2005–2015, and the 1.5 and 2 K thresholds are marked by grey horizontal lines.

Perfect cancellation of radiative imbalance by ocean carbon uptake is unlikely to occur: uptake mechanisms of carbon and heat are distinctly different in Earth system models[29,30], while the warming of the upper ocean due to removed aerosol cooling, typically neglected in idealized modelling studies[7–9,11], can lead to outgassing of CO_2 temporarily counteracting some deep-ocean carbon uptake. For these reasons, we consider perfect cancellation to be an idealization awaiting efforts to understand the fate of carbon in the Earth system.

Even including all mediating factors, there is some risk that committed warming on centennial timescales already exceeds societal aspirations to limit global warming to 1.5 to 2 degrees above pre-industrial[15]. There is 32% and 13% risk (cases d and e) that committed warming as of year 2016 already exceeds the 1.5 K warming threshold; for the 2 degree target, the risks are lower at 10% and 3%, respectively. If anthropogenic forcing keeps rising at the current rate of about $0.033\ \mathrm{W\,m^{-2}\,yr^{-1}}$ then the median value of committed warming exceeds 1.5 K in the years 2032 and 2053, for cases d and e, corresponding to additional forcing of about 0.5 and $1.2\ \mathrm{W\,m^{-2}}$. These are the estimated years by which one needs to stop all emissions to have a 50% chance of staying below 1.5 K on centennial timescales.

Our estimates of committed warming are sensitive to the relatively large uncertainty in aerosol forcing which leads to a long tail of high climate sensitivity values (Fig. 1). In the energy-balance framework, warming on centennial timescales is closely connected to TCR (Fig. 3): commitment median probability exceeds 1.5 K for TCR of 1.5 and 1.9 K for cases d and e, respectively, and exceeds 2.0 K for TCR of 2.1 and 2.6 K. This means that, if TCR of the Earth turns out to be in the upper range, then the 1.5 degree target could already be unachievable. But high values of TCR are almost exclusively associated with strong aerosol cooling ($F_{aero} < -1.0\ \mathrm{W\,m^{-2}}$, Fig. 3 red shading). Such strong aerosol cooling may be inconsistent with mid-century warming being forced[17] although the lower bound on aerosol is the subject of current debate.

On the other hand, our estimates of TCR and ECS, based as they are on the observed relationships between forcing, imbalance and temperature change, are likely to be lower than the Earth's true sensitivity for several reasons. First, observations of global-mean air surface temperature miss some of the amplified warming at high latitudes and do not carefully distinguish between surface air and water temperatures, which comprehensive models suggest may lead to slight underestimates of warming[18]. Second, forcing agents are unlikely to be equally effective in driving global temperature change. Aerosol cooling in particular may have masked more warming per unit forcing than greenhouse gas warming causes (efficacy > 1), which would act to damp estimates of sensitivity[19]. Finally, feedbacks in comprehensive models often vary with timescale or climate state, especially the pattern of surface warming[20]. In such models, the actual ECS is uniformly similar to or higher than that inferred from estimates of forcing, warming and imbalance[21]. This last issue does not seem particularly relevant: we explored the impact of time-dependent feedback (Methods and Supplementary Fig. 1) and find that, even if we choose parameters corresponding to the strongest effect found among Coupled Model Intercomparison Project Phase 5 (CMIP5) models, the increase in commitment in case d is modest (around 0.12 K). Using the model ensemble mean time dependence, the impact is only 0.04 K by the year 2100. This implies that the possible time dependence of feedbacks will have a strong impact on committed warming only if the dependencies of Earth are stronger than in any CMIP5 model.

Beyond the uncertainties introduced by relying on observations and the simple energy-balance framework the abrupt cessation of all anthropogenic emissions is such a highly idealized scenario that one might question the practical value of estimating committed warming. The reasons for doing so are partly pedagogical: committed warming defies the naive expectation that global warming stops when emissions cease and, indeed, introduces the further complications of rapid additional warming with decreasing emissions as reduced aerosol cooling unleashes masked greenhouse gas warming[3,6]. It further distinguishes future warming originating in the past from future anthropogenic emissions, which is useful for estimating the remaining headroom to exceeding target temperature thresholds. Observations-based estimates provide a conceptually transparent framework for estimating commitment that relies on a few assumptions and observables. As the Earth warms in coming decades, uncertainty in some observables will decrease, and so uncertainty in TCR will be roughly halved by year 2030[16] leading to narrowed probability for committed warming. Likewise, an improved ability to constrain aerosol forcing[17], or a deepened understanding of time- or state-dependent feedbacks can be readily implemented to narrow quantitative uncertainty on the remaining headroom to exceeding the set 1.5 and 2 degree target temperatures[15].

Methods

Methods, including statements of data availability and any associated accession codes and references, are available in the online version of this paper.

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Author contributions

The original idea for this study was conceived by T.M. R.P. and T.M. developed the methodology and wrote the manuscript.

Additional information

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Competing financial interests

The authors declare no competing financial interests.

Methods

Using an energy-balance model, two measures of the climate response to doubled CO_2 concentrations, transient climate response (TCR) and equilibrium climate sensitivity (ECS), can be estimated[22] from observations of changes between two epochs in global-mean temperature ΔT and planetary imbalance ΔQ, as well as estimates of effective radiative forcing F and the effective radiative forcing at doubled CO_2 ($F_{2\times}$) as

$$\text{TCR} = F_{2\times}\frac{\Delta T}{\Delta F} \tag{6}$$

$$\text{ECS} = F_{2\times}\frac{\Delta T}{\Delta F - \Delta Q} \tag{7}$$

where effective radiative forcing, hereafter forcing, accounts for rapid adjustments[31], that is, responses of the climate system that affect radiative balance but do not scale with temperature.

In this work the present-day epoch estimates of temperature and planetary imbalance are computed over the years 2005–2015, as coincident with the latest planetary imbalance estimate[13], while the reference epoch values are computed over the period 1859–1882; a period that, like the present-day epoch, has seen little influence of volcanic eruptions on the global energy balance[24]. Estimates of annual, global-mean surface temperature are taken from HadCRUT4 (ref. 12) with uncertainty in ΔT set to 0.08 K (ref. 24). Planetary imbalance is estimated[13] as 0.71 W m^{-2} with 5–95% confidence intervals of $\pm$0.10 W m^{-2}, and the corresponding values during the baseline period are estimated[24] at 0.15 $\pm$ 0.075 W m^{-2}. We take the forcing from greenhouse gases, aerosols and a range of other sources (ozone, stratospheric water vapour, land use, contrails, solar variability, and black carbon on snow) from Annex II of the Intergovernmental Panel on Climate Change (IPCC) Fifth Assessment[14]. The $F_{2\times}$ is set to 3.71 W m^{-2}, which is consistent with the tabulated forcing. Uncertainty in each component is also taken from the assessment report[26] and uncertainty in $F_{2\times}$ is set proportional to that of greenhouse gas forcing. The results are insensitive to the value of and amount of uncertainty in $F_{2\times}$ because ΔF appearing in the denominator is dominated by greenhouse gas forcing and so random errors roughly cancel in the estimates of TCR and ECS. Forcing is evaluated at the centre of the present-day epoch, year 2010, because the time series ends in 2011. All input to the analysis is tabulated in Supplementary Table 1.

Uncertainty in equation (7) can be represented by treating each term as a probability distribution[18,22–24]. Even symmetric uncertainty in forcing and, for ECS, in ocean heat uptake, creates skewed distributions[22] of TCR and ECS because the terms appear in the denominator. Uncertainty in temperature change, imbalance in the present-day and pre-industrial epochs, and for each component of forcing is represented by drawing random samples from a Gaussian distribution. Uncertainty in present-day forcing by greenhouse gases is assumed to be perfectly correlated with $F_{2\times}$; uncertainty in all other terms is uncorrelated by sampling each variable independently. Samples producing negative values of TCR or ECS in equation (7) are discarded as implausible but the distributions are otherwise unfiltered; this affects mean and extreme values but the median and 5–95% confidence intervals are more robust. The results are displayed in Fig. 1.

The choice of 1850–1899 as the pre-industrial reference in equations (1)–(5) is a compromise of availability of temperature observations and having boundary conditions representative of pre-industrial conditions, for example, a volcanic forcing close to the long-term mean. Yet, industrialization had already commenced during this period, and relative to 1750 there was a positive total forcing of 0.15 W m^{-2}. On the contrary, the period 2005–2015 had relatively little volcanic activity, and counting on future volcanic forcing at the level of the past forcing would yield less commitment. Together with a low efficacy of volcanic forcing[32] these two effects closely cancel.

To estimate the strength of the short-lived climate forcers (SLCFs) CH_4, NO_x and CO used in equations (3)–(5), we apply the emissions-based forcing estimates[26]. This is necessary because these are reactive species; for instance, emissions of methane lead to higher methane concentrations, as well as more tropospheric ozone and stratospheric water vapour. We can sum these forcings as follows:

$$F_{\text{SLCF}} = f_{\text{ff}} \times F_{\text{CH}_4} + F_{\text{NO}_x} + F_{\text{CO}} \tag{8}$$

whereby we assume that fossil-fuel emissions account for a fraction (f_{ff}) of anthropogenic methane emissions[33]. The result is summarized in Supplementary Table 2.

To demonstrate the applicability of TCR in estimating centennial-timescale committed warming as done in equations (4) and (5), we set up a two-layer model[34] representing the evolution of the temperature of the atmosphere, land and ocean mixed layer (T) and deep oceans (T_{d}):

$$C\frac{\partial T}{\partial t} = F + \lambda T - \epsilon\gamma(T - T_{\text{d}}) \tag{9}$$

$$C_{\text{d}}\frac{\partial T_{\text{d}}}{\partial t} = \gamma(T - T_{\text{d}})$$

with parameters chosen so as to match the median estimated ECS, TCR and the observed planetary imbalance in the early twenty-first century (Q). This was done by first setting the effective heat capacities of the atmosphere, land and ocean mixed layer (C) and deep oceans (C_{d}) to the coupled model ensemble mean from ref. 34 and the deep-ocean heat uptake efficacy parameter (ϵ) to unity to avoid time-dependent feedback. Then the feedback parameter (λ) was set to -2.07 W m^{-2} K^{-1} to obtain the inferred median ECS of 1.78 K, and finally the heat exchange coefficient (γ) was set to 1.0 W m^{-2} K^{-1} to obtain a compromise in matching the other two properties TCR $\approx$ 1.28 K and $Q \approx$ 0.70 W m^{-2} close to the respective inferred and observed values of 1.32 K and 0.71 W m^{-2} (Fig. 1 and Supplementary Table 1). Lowering γ leads to slightly larger TCR, and therefore requires a further lowering of ECS below the inferred value in order to match instrumental record warming, but then results in a smaller Q that is inconsistent with observations. Overall, though, the choice of γ within reason has little impact on modelled committed warming.

The model is run with forcing according to ref. 14 starting in the year 1750, so that the volcanic forcing is offset to obtain a long-term zero mean to avoid deep-ocean temperature drift[35]. After year 2010, the forcing is increased according to equation (4) with a linearly increasing forcing until year 2016, whereafter aerosol and SLCFs are abruptly removed. The result is displayed in Supplementary Fig. 1. Subsequently, we investigate the impact of time-dependent feedback on the transient response. We do so by setting ϵ to the CMIP5 ensemble mean value[34] of 1.28 and the maximum found among models of 1.82. In both cases we increase ECS so that the modified two-layer models match the 2005–2015 warming of the unmodified version. The temperature evolution prior to 2015 is indistinguishable among the three models. Towards the end of the twenty-first century, the impact of the multi-model mean ϵ is about 0.04 K and for the extreme case there is 0.12 K additional warming. Thus, to have an appreciable impact on estimates of committed warming, the time dependence of feedbacks of the Earth has to be well outside the range represented by CMIP5 models.

Code availability. Python scripts to conduct the calculations underlying this study and reproduce figures are archived by the Max Planck Institute for Meteorology and can be obtained by contacting either the corresponding author or publications@mpimet.mpg.de.

Data availability. HadCRUT4 data are provided by the UK Met Office Hadley Centre (http://www.metoffice.gov.uk/hadobs/hadcrut4) and forcing data are from the IPCC AR5 WG1 (http://www.climatechange2013.org/images/report/WG1AR5_AIISM_Datafiles.xlsx).

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